Subtropical Atlantic salinity variability and Atlantic meridional circulation during the last deglaciation

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ABSTRACT

During the last deglaciation (ca. 21–10 ka), freshening of the North Atlantic surface likely caused reductions in Atlantic meridional overturning circulation (AMOC); the mechanisms related to AMOC recovery remain poorly understood. Here we present three new deglacial surface temperature and $\delta^{18}O_{seawater}$ ($\delta^{18}O_{sw}$) reconstructions from the western subtropical North and South Atlantic. Similarities to tropical Caribbean and western Atlantic $\delta^{18}O_{sw}$ records suggest that a salty surface water mass accumulated in the western Atlantic from 27°S to 33°N during periods of reduced AMOC. However, $\delta^{18}O_{sw}$ decreases led deep AMOC resumption by hundreds of years. We suggest that the northward export of salt previously trapped in the western Atlantic resulted in the early establishment of a shallow overturning circulation that eventually culminated in deep AMOC resumption, implying that AMOC may constitute a self-limiting system.

INTRODUCTION

The tropics and subtropics of the North Atlantic are regions of net evaporation, which increases the salinity of surface waters in the subtropical North Atlantic. Subduction of salty northeastern subtropical waters into the thermocline also increases the salinity of western Atlantic thermocline waters. These high-salinity waters are an important component of the upper limb of modern Atlantic meridional overturning circulation (AMOC) (Broecker et al., 1990; Schmitz, 1995), hence changes in their salinity may have influenced AMOC in the past (Schmidt et al., 2004, 2006; Weldeab et al., 2006). Once these waters reach the subpolar North Atlantic and cool, convection occurs, forming North Atlantic deepwater and continuing the AMOC (Schmitz, 1995). During the Last Glacial Maximum (LGM, ca. 21 ka), AMOC strength was apparently reduced relative to present, and during the subsequent deglaciation underwent two oscillations in strength with attendant effects on climate (Boyle and Keigwin, 1987; McManus et al., 2004; Robinson et al., 2005). The larger of these reductions spanning the Oldest Dryas cold event (ca. 18.5–14.7 ka) was likely in response to the input of freshwater from initial retreat of Northern Hemisphere ice sheets ca. 19 ka (Clark et al., 2004), reinforced by the melting of icebergs during Heinrich Event 1 (H1) (17.5–16.0 ka) (McManus et al., 2004). A smaller reduction occurred during the Younger Dryas cold event (ca. 13–11.5 ka), probably due to the routing of North American freshwater to the northern North Atlantic (e.g., Broecker et al., 1990; Carlson et al., 2007) and increased iceberg discharge (McManus et al., 2004). Following each weakening, the AMOC eventually resumed (McManus et al., 2004). However, the mechanisms behind these resumptions are unclear (Weaver et al., 2003; Knorr and Lohmann, 2003; Schmidt et al., 2004; Weldeab et al., 2006; Leduc et al., 2007), especially in the case of the Oldest Dryas, when the deep AMOC (>2500 m water depth) resumption lagged the end of H1 by ~1.3 k.y. (McManus et al., 2004; Robinson et al., 2005). Because the subtropical western Atlantic surface hydrology exerts a strong control on modern AMOC, we evaluate the possibility that the subtropics played an important role in modulating AMOC in the past.

METHODS

We reconstructed calcification temperature (CT) and sea-surface salinity (SSS)–dependent $\delta^{18}O_{seawater}$ ($\delta^{18}O_{sw}$) from three cores located along the northward return flow of surface water associated with AMOC (Broecker et al., 1990; Schmitz, 1995). Core OCE326-GGC5 is located on the Bermuda Rise, KNR140–51GGC is on the Blake Outer Ridge, and KNR159–5–36GGC is along the Brazilian margin (Fig. 1). For the Bermuda Rise, we measured $\delta^{18}O_{calcite}$ and Mg/Ca of the planktonic foraminifera *Globorotalia inflata* (thermocline dwelling; Anand et al., 2003) for the core top, Holocene, and deglacial sections of the core (Fig. 2; GSA Data Repository Fig. DR1[1]). For the Blake Outer Ridge and the Brazilian margin, we measured $\delta^{18}O_{calcite}$ and Mg/Ca of the planktonic foraminifera *Globigerinoides ruber* (white) (surface dwelling; Anand et al., 2003) for the core tops and deglacial sections of the two cores (Fig. 2 and Fig. DR1). After converting Mg/Ca to CT, we corrected $\delta^{18}O_{calcite}$ for changes in CT and continental ice volume to reconstruct $\delta^{18}O_{sw}$ (Fig. 3 and Fig. DR1). In the case of the Bermuda Rise with its deep water depth, we developed a weight-dependent Mg/Ca-CT calibration to account for the effects of dissolution. Detailed methods and reservoir corrected-calibrated age model construction are provided in the GSA Data Repository. We compare our $\delta^{18}O_{sw}$ results to simulations using the National Aeronautics and Space Administration Goddard Institute for Space Studies fully coupled atmosphere-ocean general circulation model (GCM) ModelE-R that tracks water isotopes (Schmidt et al., 2007). ModelE-R was forced with 0.1 Sverdrups (1 Sverdrup [Sv] = 10^6 m^3 s^{-1}) of fresh water to the North Atlantic for 100 yr. We compare the average of the past 20 yr of this experiment to the control simulation.

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[1]GSA Data Repository item 2008246, detailed methods and results description, core data, and additional tables and figures, is available online at www.geosociety.org/pubs/ft2008.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

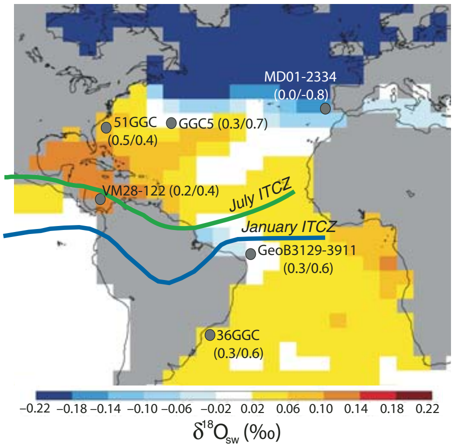

Figure 1. Locations of cores GGC5 on Bermuda Rise (33°42′N, 57°35′W, 4550 m water depth), 51GGC on Blake Outer Ridge (32°47′N, 76°17′W, 1790 m water depth), 36GGC on Brazilian margin (27°31′S, 46°28′W, 1268 m water depth), VM28–122 (11°34′N, 78°25′W, 3623 m water depth) in Caribbean (Schmidt et al., 2004), GeoB3129–3911 (4°37′S, 36°38′W, 830 m water depth) on northeastern Brazil margin (Weldeab et al., 2006), and MD01–2334 (37°33′ N, 10°08′W, 2460 m water depth) on Iberian margin (Skinner and Shackleton, 2006). Green line is modern July Intertropical Convergence Zone (ITCZ) location; blue line is January location. Also shown is National Aeronautics and Space Administration Goddard Institute for Space Studies general circulation ModelE-R (Schmidt et al., 2007) surface $\delta^{18}O_{sw}$ response to ~50% reduction in Atlantic meridional overturning circulation (AMOC) strength. Changes in core $\delta^{18}O_{sw}$ during reduced AMOC are indicated in parentheses (first value is for Younger Dryas, second is for Oldest Dryas). Because $\delta^{18}O_{sw}$ changes differ somewhat in timing between records, changes in core $\delta^{18}O_{sw}$ were calculated using average of three points prior to $\delta^{18}O_{sw}$ change and average of three points just after change.

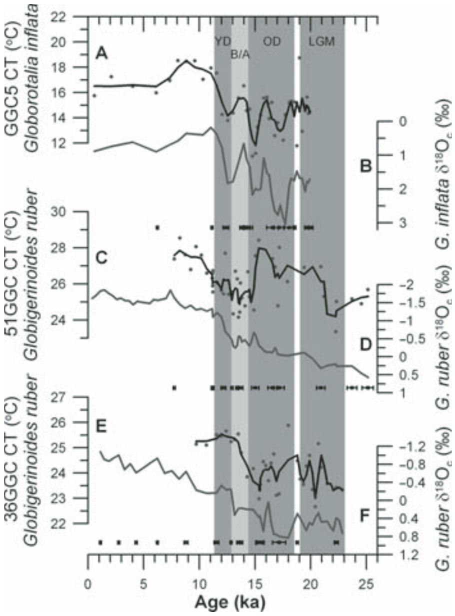

Figure 2. Subtropical calcification temperature (CT) (gray dots with 3 point smoothed black line) and $\delta^{18}O_{calcite}$ ($\delta^{18}O_c$) (gray line). A: Bermuda Rise (GGC5) *Globorotalia inflata* CT. B: GGC5 $\delta^{18}O_c$. C: Blake Outer Ridge (51GGC) *Globigerinoides ruber* CT. D: 51GGC *G. ruber* $\delta^{18}O_c$. E: Brazilian margin (36GGC) *G. ruber* CT. F: 36GGC *G. ruber* $\delta^{18}O_c$. Black boxes denote radiocarbon age control. Dark gray bars denote Last Glacial Maximum (LGM), and Oldest (OD) and Younger Dryas (YD) cold events. Light gray bar denotes Bølling/Allerød Warm Period (B/A).

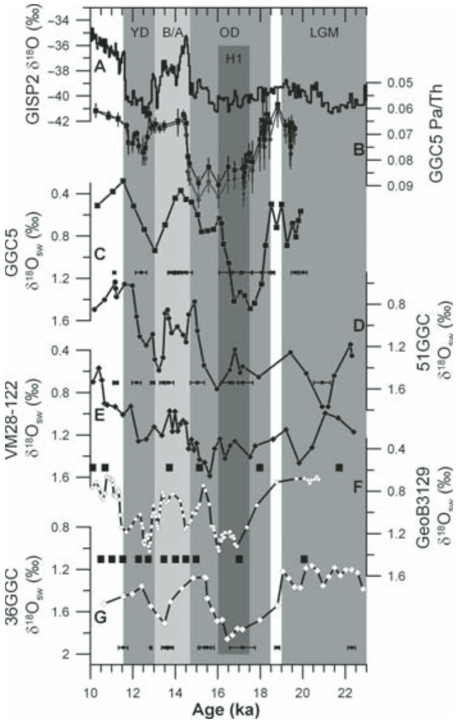

Figure 3. Subtropical-tropical sea-surface salinities (SSS) ($\delta^{18}O_{sw}$), Atlantic meridional overturning circulation (AMOC), and North Atlantic climate. A: Summit Greenland (GISP—Greenland Ice Sheet Project) $\delta^{18}O$ (Grootes et al., 1993). B: Bermuda Rise $^{231}Pa/^{230}Th$ record from subtropical North Atlantic (McManus et al., 2004). C: $\delta^{18}O_{sw}$ from Bermuda Rise (square symbols, 3 point smoothing). D: $\delta^{18}O_{sw}$ from Blake Outer Ridge (circular symbols, 3 point smoothed). E: $\delta^{18}O_{sw}$ from Caribbean (diamond symbols) (Schmidt et al., 2004). F: $\delta^{18}O_{sw}$ from northeastern Brazil margin (open circular symbols, 3 point smoothed) (Weldeab et al., 2006). G: $\delta^{18}O_{sw}$ from Brazilian margin (open diamond symbols, 3 point smoothed). Black boxes represent radiocarbon age control. Records are smoothed to reduce significance of noisy data while increasing significance of the signal (Bevington and Robinson, 2002). Medium gray bars denote Last Glacial Maximum (LGM), and Oldest (OD) and Younger Dryas (YD) cold events. Dark gray bar denotes timing of Heinrich Event 1 (H1). Light gray bar denotes Bølling/Allerød Warm Period (B/A).

RESULTS

At the LGM, the Blake Outer Ridge, Brazilian margin, and Bermuda Rise CT were ~2.5, 1, and 2.5 °C colder than modern CT, respectively, in agreement with CLIMAP (climate: long-range investigation, mapping, and prediction) estimates (CLIMAP Project Members, 1981) (Fig. 2). During the LGM, the Blake Outer Ridge CT increased by ~1 °C with a second increase of ~1.5 °C between ca. 17 and 15.4 ka. CT subsequently cooled by ~2 °C, before warming again ca. 13 ka. The Brazilian margin CT remained relatively constant following the LGM until 15–14 ka, when it began a gradual deglacial warming trend. In contrast, the Bermuda Rise CT record shows 1–3 °C cooling to minima ca. 17, 15, and 13 ka, with 2–3 °C of warming ca. 16.5, 14.7, and 12 ka.

During the last deglaciation, the three subtropical $\delta^{18}O_{sw}$ records show similar, but not identical, millennial-scale variability (Figs. 3C, 3D, and 3G). Bermuda Rise and Brazilian margin $\delta^{18}O_{sw}$ began to increase at 18.5–18.0 ka, reaching maxima ca. 16.5 ka, with subsequent decreases. In contrast, Blake Outer Ridge $\delta^{18}O_{sw}$ was highest during the LGM with a subsequent decrease. It began a gradual increase again 18.5–18.0 ka, reaching a maximum ca. 16 ka. $\delta^{18}O_{sw}$ at all three sites decreased abruptly between 15.5 and 15.2 ka. At 13–13.5 ka, all three sites exhibit another maximum in $\delta^{18}O_{sw}$. Assuming modern $\delta^{18}O_{sw}$-SSS relationships at the core sites (LeGrande and Schmidt, 2006), these millennial-scale $\delta^{18}O_{sw}$ events reflect SSS variations of ≥1 practical salinity unit (psu).

DISCUSSION AND CONCLUSIONS

The three CT records from the western subtropical Atlantic show distinctly different temperature trends across the last deglaciation. At the Bermuda Rise, CT cooled twice during the Oldest Dryas and once during the Younger Dryas, periods of reduced AMOC strength ($^{231}Pa/^{230}Th$ measured from the same core) (McManus et al., 2004). These Bermuda Rise CT decreases are similar in both timing and magnitude to a CT record off the Iberian margin, possibly reflecting the influence of subducted cooler thermocline waters from the eastern subtropical North Atlantic (Skinner and Shackleton, 2006). While there is not a coincident oscillation in $^{231}Pa/^{230}Th$ ca. 16.5 ka, when CT increased at the Bermuda Rise (Fig. 2), we note that a similar Iberian margin CT oscillation and decreased North Atlantic bottom-water nutrient content ca. 16.5 ka (Boyle and Keigwin, 1987; Skinner and Shackleton, 2006) suggest that the Bermuda Rise CT oscillation may be more than a local phenomenon.

The relatively warm, close to modern CT at the Blake Outer Ridge following the LGM and through the Oldest Dryas suggests the accumulation of heat during reduced AMOC, with the subsequent decrease implying the resumption of northward heat transport (Fig. 2). We note that while this may be a local phenomenon, deglacial pollen records from Florida suggest warming across the Oldest Dryas (Grimm et al., 2006) and CT warming from 24 to 26–27 °C is observed at Blake Outer Ridge during H5 and H5a (Schmidt et al., 2006). Warming of 1–2 °C is also observed in the tropical western Atlantic and Caribbean during the Oldest Dryas (Rühlemann et al., 1999; Schmidt et al., 2004; Weldeab et al., 2006) (Fig. 1). Although this warming does not occur at precisely the same time in all cores (perhaps to some extent due to age uncertainties and/or local effects), the results generally suggest that heat was stored in the western tropical and possibly northwestern subtropical Atlantic during intervals of reduced AMOC.

The new $\delta^{18}O_{sw}$ records bear a similarity to AMOC records (Boyle and Keigwin, 1987; McManus et al., 2004; Robinson et al., 2005), with decreased AMOC during the Oldest and Younger Dryas generally associated with increased subtropical western to central Atlantic $\delta^{18}O_{sw}$, and presumably salinity (Fig. 3). This relationship has been documented in the Caribbean (Schmidt et al., 2004) and western tropical Atlantic (Weldeab et al., 2006) during the last deglaciation (Figs. 1 and 3), and during marine isotope stage 3 (ca. 45–60 ka) at the Blake Outer Ridge (Schmidt et al., 2006). While the timing of $\delta^{18}O_{sw}$ changes differs among the core sites (possibly reflecting age uncertainties and/or local effects), all records suggest that $\delta^{18}O_{sw}$ started to increase 18–19 ka (Fig. 3). At the Brazilian margin, the Bermuda Rise, and the western tropical Atlantic, $\delta^{18}O_{sw}$ increases were relatively abrupt. At the Blake Outer Ridge and in the Caribbean, $\delta^{18}O_{sw}$ initially decreased from minimum values ca. 21 and 20 ka, respectively, and gradually increased through the Oldest Dryas. Following an interval of high $\delta^{18}O_{sw}$, the Bermuda Rise shows an early decrease ca. 16.5 ka that is not seen in the other records. This suggests the arrival of fresher water that partly balanced salt accumulation, implying the influence of thermocline water sourced from the fresher northeastern subtropical Atlantic (Fig. 1) (Skinner and Shackleton, 2006). The other surface records indicate a later decrease of $\delta^{18}O_{sw}$ values at 16–15.2 ka that is also registered as a smaller decrease at the Bermuda Rise.

A comparison of $\delta^{18}O_{sw}$ and $^{231}Pa/^{230}Th$, both from the Bermuda Rise, suggests an ~0.5 k.y. lead of the $\delta^{18}O_{sw}$ decrease ca. 15.2 ka over deep AMOC resumption ca. 14.7 ka. This lead is also observed in the Blake Outer Ridge (0.2–0.7 k.y.), the Brazilian margin (0.5–0.7 k.y.), and the Caribbean (0.5–0.7 k.y.) (lead range reflects calibrated ^{14}C age 2σ) (Fig. 3). The western tropical Atlantic record (GeoB3129) suggests an earlier decrease in $\delta^{18}O_{sw}$ than the other four records, but this portion of GeoB3129 has no radiocarbon dates, and the timing of the decrease overlaps with the other records within age model uncertainties (Weldeab et al., 2006). Thus, our new records suggest that similar deglacial changes in $\delta^{18}O_{sw}$ occurred in the western tropical and subtropical Atlantic along the AMOC return flow path from 27°S to 33°N.

The SSS proxy and $\delta^{18}O_{sw}$ records from the subpolar North Atlantic indicate decreased $\delta^{18}O_{sw}$ and SSS during the Oldest and Younger Dryas (e.g., Clark et al., 2004; Skinner and Shackleton, 2006; Carlson et al., 2007), suggesting that during periods of reduced AMOC, salty water accumulates in the central to western subtropical and tropical Atlantic while freshwater influences the subpolar regions and northeastern subtropics. However, the $\delta^{18}O_{sw}$ response to reduced AMOC during the Younger Dryas is smaller and more limited in extent than the Oldest Dryas response, with the largest Younger Dryas signals observed in the tropical and northwestern subtropical Atlantic (Fig. 1). This smaller response may reflect the smaller magnitude and shorter lived AMOC reduction during the Younger Dryas relative to the Oldest Dryas (McManus et al., 2004; Robinson et al., 2005), and thus less salt accumulation.

To test the hypothesis that deglacial western Atlantic $\delta^{18}O_{sw}$ oscillations are controlled by AMOC, we use results from the idealized ModelE-R 0.1 Sv freshwater forcing experiment conducted for Paleoclimate Modeling Intercomparison Project Phase II (PMIP2) (Stouffer et al., 2006), but now including $\delta^{18}O$ tracers. This experiment is run from the preindustrial simulation and does not include full deglacial boundary conditions (e.g., ice sheets, ice sheet runoff, sea level, orbital forcing, greenhouse gases, salinity), limiting model-data comparison only to the spatial $\delta^{18}O_{sw}$ pattern from a reduction in AMOC. The simulation shows an ~50 % reduction in AMOC, similar to "hosing" experiments with other GCMs (Stouffer et al., 2006). Surface $\delta^{18}O_{sw}$ increases in the subtropical and tropical western to central Atlantic (Fig. 1) due to reduced northward advection of the surface water mass in response to this AMOC reduction (LeGrande and Schmidt, 2008). While the magnitudes of modeled $\delta^{18}O_{sw}$ increases are less than those observed in paleorecords, possibly due to the lack of accurate deglacial boundary conditions or the duration and/or amount of the AMOC reduction, the sign and spatial distribution of the $\delta^{18}O_{sw}$ response are in agreement with the $\delta^{18}O_{sw}$ records (Fig. 1). Thus the covariation of western Atlantic surface and subsurface $\delta^{18}O_{sw}$ with AMOC suggests that millennial-scale variations in AMOC exerted a major control on subtropical and tropical western Atlantic SSS through its role in alternatively trapping and advecting salt.

Other studies have attributed deglacial subtropical and tropical western Atlantic SSS variability to migrations in the Intertropical Convergence Zone (ITCZ) (Fig. 1) (Schmidt et al., 2004, 2006; Weldeab et al., 2006; Leduc et al., 2007). Southward ITCZ displacement during North Atlantic cooling and reduced AMOC may enhance the influence of AMOC on SSS in the northern subtropics and tropics through increased evaporation (Peterson et al., 2000; Lea et al., 2003; Schmidt et al., 2004, 2006; Weldeab et al., 2006; Benway et al., 2006). Or, it may act as a positive feedback in the southwestern tropical Atlantic because of orographically blocked moisture transport from the Atlantic to the Pacific with a southward ITCZ (Wang et al., 2004; Leduc et al., 2007). However, both data and model results suggest that while ITCZ migration may cause large changes in seasonal precipitation, the precipitation-driven SSS anomalies are tempered by mixing during transport with ocean currents, in agreement with a previous study of Holocene Pacific salinity changes (Oppo et al., 2007). The GCM simulation shows a southward shift in ITCZ coinciding with increased precipitation and decreased $\delta^{18}O_{sw}$ of Amazon River discharge, but at least along the western to central subtropical and tropical Atlantic return flow path, diminished salt export from the subtropics and tropics due to reduced AMOC overwhelms the ITCZ effect (Fig. 1 and Fig. DR2).

At the end of the Younger Dryas, AMOC increased coincident with the reduction in freshwater runoff to the North Atlantic (Carlson et al., 2007). However, deep AMOC and deep-ocean ventilation did not immediately resume after the end of H1 ca. 16 ka, but rather lagged until ca. 14.7 ka (Fig. 3) (McManus et al., 2004; Robinson et al., 2005). Southern Ocean sources and increased leakage of the Agulhas Current have been suggested for the turn-on of AMOC by some GCMs and faunal reconstructions (Knorr and Lohmann, 2003; Weaver et al., 2003; Peeters et al., 2004), with increased tropical-subtropical salinity preconditioning the North Atlantic and acting as a positive feedback after the initial triggering of AMOC (Schmidt et al., 2004, 2006; Weldeab et al., 2006).

In contrast, our results indicate a 0.2–0.7 k.y. lead of $\delta^{18}O_{sw}$ decreases over deep AMOC resumption and deep-ocean ventilation ca. 14.7 ka, with the initial decrease beginning shortly after the end of freshwater forcing during H1 (Fig. 3). Decreased ventilation ages of shallower (<2500 m) North Atlantic waters and a decrease in North Atlantic bottom-water nutrients also occurred several hundred years prior to the onset of deep-ocean ventilation ca. 14.7 ka (Boyle and Keigwin, 1987; Robinson et al., 2005). Together, these AMOC proxies and the $\delta^{18}O_{sw}$ records suggest an invigoration of shallow overturning with northward salt advection and shifting bottom-water mass boundaries hundreds of years prior to the increase in deep AMOC, that may not be detected in the $^{231}Pa/^{230}Th$ (Thomas et al., 2006). This shallow overturning could be the gradual resumption of AMOC from a stagnant state following the end of freshwater forcing, which is simulated by fully coupled GCMs (Stouffer et al., 2006), and is possibly caused by the buildup of a large enough SSS gradient (~1 psu) in the tropics and subtropics (Broecker et al., 1990). The attendant northward heat transport may be reflected in Summit Greenland warming prior to the Bølling (Fig. 3). Northward heat transport could also initiate sea ice retreat that culminates in deep AMOC resumption upon the release of heat trapped beneath the sea ice with further atmospheric warming (Kaspi et al., 2004), explaining the large Summit Greenland temperature increase ca. 14.7 ka. Thus AMOC may have an internal negative feedback that limits AMOC reductions, or a salt oscillator (Broecker et al., 1990), constituting a self-limiting system.

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